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Weak overturning circulation and high Southern Ocean nutrient utilization maximized glacial ocean carbon ☆

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ABSTRACT

Circulation changes have been suggested to play an important role in the sequestration of atmospheric CO_2 in the glacial ocean. However, previous studies have resulted in contradictory results regarding the strength of the Atlantic Meridional Overturning Circulation (AMOC) and three-dimensional, quantitative reconstructions of the glacial ocean constrained by multiple proxies remain scarce. Here we simulate the modern and glacial ocean using a coupled physical-biogeochemical, global, three-dimensional model constrained simultaneously by $\delta^{13}C$, radiocarbon, and $\delta^{15}N$ to explore the effects of AMOC differences and Southern Ocean iron fertilization on the distributions of these isotopes and ocean carbon storage. We show that $\delta^{13}C$ and radiocarbon data sparsely sampled at the locations of existing glacial sediment cores can be used to reconstruct the modern AMOC accurately. Applying this method to the glacial ocean we find that a surprisingly weak (6–9 Sv or about half of today's) and shallow AMOC maximizes carbon storage and best reproduces the sediment isotope data. Increasing the atmospheric soluble iron flux in the model's Southern Ocean intensifies export production, carbon storage, and further improves agreement with $\delta^{13}C$ and $\delta^{15}N$ reconstructions. Our best fitting simulation is a significant improvement compared with previous studies, and suggests that both circulation and export production changes were necessary to maximize carbon storage in the glacial ocean.

1. Introduction

During the Last Glacial Maximum (LGM, ~ 21 ky before present) atmospheric pCO_2 was about 100 ppm lower than its pre-industrial (PI) value (Marcott et al., 2014). Reconstructions of stable carbon isotope ($\delta^{13}C$) distributions from LGM sediments indicate a shallower Atlantic Meridional Overturning Circulation (AMOC) and more remineralized nutrients and carbon in the deep Atlantic (Gebbie, 2014). However, no consensus on the glacial AMOC strength exists, with studies inferring weaker (Lynch-Stieglitz et al., 2007), similar (Böhm et al., 2015), or stronger (Kurahashi-Nakamura et al., 2017) overturning rates. Radiocarbon ($\Delta^{14}C$) reconstructions suggest that the deep ocean was more isolated from the atmosphere, which has been suggested to explain the full glacial–interglacial change in atmospheric CO_2 (Sigman et al., 2010; Sarnthein et al., 2013).

On the other hand, an increase in biological production and export of organic matter could have transferred a substantial amount of carbon from the surface and atmosphere to deep waters during the LGM (Martin, 1990). In most of the present Southern Ocean (SO) productivity and macro-nutrient (nitrate, phosphate) utilization is limited by low concentrations of dissolved iron (DFe). Higher dust deposition could have fertilized this region with iron and increased its efficiency as a carbon sink during the LGM. Observations indicate that atmospheric dust fluxes into the SO (Maher et al., 2010; Lambert et al., 2015) and soluble iron content in dust (Conway et al., 2015) were increased in the LGM compared to today, and nutrient utilization (inferred from $^{15}N/^{14}N$) was more efficient (Martínez-García et al., 2014). However, the effects of increased SO nutrient utilization and changes in deep ocean circulation on three-dimensional glacial ocean carbon storage remain unquantified.

The goal of this work is to use a model-data comparison to constrain the state of the LGM ocean, including its AMOC, SO nutrient utilization and to separately quantify their effects on ocean carbon storage. We use a three-dimensional, global climate/circulation/biogeochemistry model (Muglia et al., 2017), that includes, for the first time, three key isotopes (^{14}C, ^{13}C, ^{15}N) simultaneously, as well as interactive iron cycling. Previous studies did not use interactive iron cycling (Schmittner and Somes, 2016) or only used

☆ Fully documented templates are available in the elsarticle package on CTAN.

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individual isotopes (mostly $\delta^{13}C$, Brovkin et al., 2007; Tagliabue et al., 2009; Bouttes et al., 2011; Menviel et al., 2017), which did not allow separation between circulation and export production effects in model/data comparisons. Our comparisons of multiple simulated isotopes with sedimentary reconstructions exploit complementary constraints provided by these isotopes (Schmittner and Somes, 2016) and offer novel insights into the glacial ocean.

A detailed analysis of the different components of the ocean's carbon cycle and their effects on atmospheric CO_2 is beyond the scope of this paper and will be presented elsewhere (Khatiwala et al., in preparation).

2. Methods

2.1. Physical and biogeochemical model

We use the global ocean circulation model from the University of Victoria (UVic) (Weaver et al., 2001), version 2.9. It consists of a three-dimensional dynamical ocean with 19 vertical levels at $3.6° \times 1.8°$ horizontal resolution governed by the primitive equations, coupled to a two-dimensional single-level atmosphere, with moisture and heat balances and fluxes between the two mediums, and a dynamical sea ice model. The model is coupled to a dynamic land vegetation model (Meissner et al., 2003). For the pre-industrial simulation wind stress, winds used in air–sea fluxes and horizontal moisture and heat advection, and clouds are prescribed from a present-day monthly climatology (Kalnay et al., 1996). Evaporation, precipitation, air moisture and temperature are calculated by the two-dimensional model. Background vertical diffusivity was set to 3.5×10^{-5} m^2/s. Our diapycnal mixing scheme includes a three-dimensional parametrization of tide effects (Schmittner and Egbert, 2013), but it does not consider changes in tidal energy dissipation. Isopycnal eddy diffusivity was set to 1.2×10^3 m^2/s. Higher values are applied in the tropics, to include effects of the Equatorial Intermediate Current System on temperature, salinity and dissolved oxygen (Getzlaff and Dietze, 2013).

LGM runs use an atmospheric CO_2 value of 185 ppm, and orbital parameters corresponding to 21 kyr. Using realistic atmospheric CO_2 is important for an unbiased model-data comparison because fractionation of ^{13}C during photosynthesis (Schmittner et al., 2013) and radiocarbon ages (Galbraith et al., 2015) depend on atmospheric CO_2. The wind stress fields include a multi-model mean LGM anomaly from the Paleoclimate Model Intercomparison Project Phase 3 (PMIP3), which are largest over the north Atlantic Ocean due to the presence of the Laurentide Ice Sheet in North America and tend to intensify the AMOC (Muglia and Schmittner, 2015). For the LGM continental ice sheets, we use the reconstruction from the PMIP3 set up (Abe-Ouchi et al., 2015). We also apply a global 1 PSU addition to salinity to account for sea level drop. Changes in river routings were not included. The model's land-sea mask and the resolved bathymetry were not changed from preindustrial conditions.

In LGM runs, the fixed atmospheric CO_2 of 185 ppm forces the ocean to equilibrate its carbon to a lower pre-formed value than in preindustrial runs that use 280 ppm. Thus the global ocean DIC content is lower in LGM than in preindustrial simulations. It is not a goal of this work to infer glacial–interglacial changes in atmospheric CO_2 concentrations, but to compare different modeled LGM configurations with proxy observations, and also to assess their DIC content and distribution. The fact that none of the LGM simulations has a larger DIC inventory than the preindustrial indicates that none of the simulations would result in LGM levels of atmospheric CO_2 in interactive simulations with prognostic atmospheric CO_2 and a constant carbon inventory in the ocean–atmosphere–land system.

UVic is coupled with the Model of Ocean Biogeochemistry and Isotopes (MOBI), version 1.8. It includes prognostic equations for PO_4, NO_3, DFe, O_2, DIC, dissolved organic matter, particulate organic matter (detritus), particulate iron, phytoplankton, zooplankton and diazotrophs. It is equipped with an advection–diffusion scheme, so tracers depend on ocean physics as well as biogeochemical interactions. The P content of the ocean is assumed constant, but the N cycle includes N_2 fixation by diazotrophs (source of NO_3), and water column and benthic denitrification (sinks of NO_3) (Somes and Oschlies, 2015).

The iron cycle includes external inputs from atmospheric dust deposition, sedimentary release, and hydrothermal fluxes (Muglia et al., 2017). Atmospheric deposition is calculated from a prescribed surface field and added to DFe at the surface. Sedimentary release is proportional to the flux of organic matter reaching the ocean floor, and to a sub-grid bathymetry parameter used to account for unresolved bottom features. Bulk hydrothermal iron fluxes at mid-ocean ridges are added locally to DFe concentrations at the corresponding grid boxes.

2.2. Moisture transport and circulation

Various processes and model parameters can affect the meridional overturning circulation (MOC). Here we modify the meridional southern hemisphere moisture transport (F_{qSH}), which has been proven to be an efficient way to affect Antarctic Bottom Water (AABW) production and AMOC strength, producing different MOC states. A decreased F_{qSH} leads to saltier AABW and a weaker AMOC (Saenko et al., 2003). It has been suggested that changes in atmospheric water vapor transport played a role in glacial–interglacial changes in SO stratification and CO_2 (Sigman et al., 2007), but estimates of this transport in the LGM are uncertain.

Specific humidity in the UVic model is controlled by the equation

$$\rho_a H\left(\frac{\partial q}{\partial t} + \nabla(\mathbf{u}q) - \nabla(\mu\nabla q)\right) = \rho_o(E - P), \tag{1}$$

where ρ_a and ρ_o are the density of air and water, respectively, $H = 1.8$ km is a scale height, $\mathbf{u}$ is the mean wind velocity for the advection of moisture, E and P are evaporation and precipitation fluxes, respectively, q is the specific humidity, and μ is the moisture eddy diffusivity. μ accounts for the transport of moisture due to any process that cannot be explained by the resolved advection. It has a trigonometric dependence on latitude (Fig. S1), with higher values at mid latitudes, where eddies play an important role in atmospheric circulation. Larger diffusivities in the southern hemisphere compared with the northern hemisphere improve agreement with modern observations (Saenko et al., 2003), so an anomaly μ_{SH} is added to the moisture diffusivity in the southern hemisphere.

Variations of μ_{SH} produce changes in the meridional transport of humidity from the equator to the SO (Fig. S1), and control the buoyancy of Antarctic Circumpolar waters. Hence we can vary μ_{SH} among experiments and test the effect of different deep circulations and water mass distributions on biogeochemistry, isotopes, radiocarbon ages and carbon storage. We do this by multiplying μ_{SH} by a factor ϵ_q that goes from 1 (default μ_{SH}) to -0.5.

2.3. Iron fluxes

PI and LGM surface soluble iron fluxes were calculated from dust flux monthly climatologies (Lambert et al., 2015, see Muglia et al., 2017 for more details). Compared to the PI, the calculated LGM surface soluble iron fluxes exhibit an increase in parts of the SO, to the east of the Patagonian region and around continents (Fig. S2). Yearly atmospheric surface fluxes to the ocean south of 35°S